

Building Icelandic Igneous Crust by Repeated Melt Injections

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Key Points

- Repeating earthquakes are detected and located down to a magnitude of ML-0.5
- Opening tensile cracks produce earthquakes as melt intrudes the Icelandic crust
- Multiple igneous sills were intruded into the lower crust at 20 km depth.

Abstract

Observations of microseismicity provide a powerful tool for mapping the movement of melt in the crust. Here we record remarkable sequences of earthquakes 20 km below the surface in the normally ductile crust in the vicinity of Askja volcano, in north-east Iceland. The earthquakes occur in swarms consisting of identical waveforms repeating as frequently as every 8 s for up to 3 hours. We use template waveforms from each swarm to detect and locate earthquakes with an automated cross-correlation technique. Events are located in the lower crust and are inferred to be the result of melt being injected into the crust. During melt intrusion high strain rates are produced in conjunction with high pore-fluid pressures from the melt or exsolved carbon dioxide. These cause brittle failure on high angle fault planes located at the tips of sills. Moment tensor solutions show that most of the earthquakes are opening cracks accompanied by volumetric increases. This is consistent with the failure causing the earthquakes by melt injection opening new tensile

cracks. Analysis of the magnitude distribution of earthquakes within a swarm reveals a complicated relationship between the imposed strain rates and the fluids that cause brittle failure. The magnitude of the earthquakes is controlled by the distance fluids can migrate along a fault, whereas the frequency of the events is controlled by the strain rate. Faults at the tips of sills act to focus melt transport between sills and so must be an important method of transporting melt through the lower crust.

Index Terms

8416 Mid-oceanic ridge processes

7230 Seismicity and tectonics

7280 Volcano seismology

8434 Magma migration and fragmentation

Keywords

earthquake, fluids, intrusion, melt injection, moment tensor solutions, non-double-couple sources

1. Introduction

New crust at igneous rifts is generated by the injection of magma into the crust. Only a small proportion of the injected melt is erupted at the surface, with the remainder fractionating and freezing at depth. Typically around 70 – 80% of the melt freezes in the crust without reaching the surface [Kelemen *et al.*, 1997; White *et al.*, 2008]. Geochemical analysis of melt inclusions in erupted basalts in Iceland suggest that the melt stalls and fractionated in one or more sills or magma chambers within the crust before it is erupted [MacLennan *et al.*, 2001]. Direct observations of igneous sills intruding the lower crust come from seismic reflection profiles across highly stretched crust on the early Tertiary continent-ocean boundary near the Faroe Islands in the North Atlantic [White *et al.*, 2008; Roberts *et al.*, 2009]. This is an analogue of the current rift under Iceland because it was generated by stretching above a mantle plume. Lower crustal sills have also been imaged seismically at the Juan de Fuca Ridge spreading center [Canales *et al.*, 2009] and the East Pacific

Rise [Marjanovi *et al.*, 2014]. Exposed Moho sections in ophiolites, which represent former oceanic crust, often exhibit interlayered mafic-ultramafic sections caused by sills in the vicinity of the Moho and in the lower crust [Boudier and Nicholas, 1995].

Indirect evidence for melt intrusion into the crust in volcanic regions can often be deduced from surface deformation recorded by GPS and InSAR techniques. Generally these methods have increasingly poor resolution with increasing depth of subsurface deformation, although they provide good constraints on the intrusion of melt into shallow magma chambers under active volcanoes [Sigmundsson *et al.*, 2010; Bagnardi and Amelung, 2012; Masterlark *et al.*, 2012]. Regions where melt is stored may also be mapped from the effect of the melt on physical properties such as conductivity using magneto-telluric methods, or compressional and shear wave velocities using seismic tomography [Toomey *et al.*, 1990; Dunn *et al.*, 2000; Mitchell *et al.*, 2013; Lin *et al.*, 2014]. All these methods generally lack resolution both spatially and temporally in the lower crust and cannot show details of how melt is stored or how it is transported from one place to another.

Microseismicity provides a powerful high-resolution method for directly recording regions where melt is moving in, or through the crust. For example, evidence of melt evacuation from a series of stacked sills in the crust and the upper mantle has been inferred from seismicity recorded to depths of 30 km beneath Eyjafjallajökull in southeast Iceland during its last eruption in 2010 [Tarasewicz *et al.*, 2012]. Note, however, that because high-frequency microearthquakes attributed to the movement of melt are caused by brittle failure of the rock, they record the propagation of new cracks, the fracture of melt frozen in a dyke or sill, or fracturing around the margins of an intrusion as it deforms the country rock. They do not directly record the flow of melt.

In Iceland, the brittle-ductile boundary is at 7–9 km depth, and under normal circumstances earthquakes are not recorded from the ductile lower crust because geological strain rates are too low to cause brittle failure. However, melt movement may generate strain rates and pore pressures that are sufficiently high to cause brittle failure and therefore seismicity [Webb and Dingwell, 1990; White *et al.*, 2011]. High strain rates and resultant seismicity can also be produced by the exsolution of volatiles from melt. Carbon dioxide exsolves

from basalts at depths of 15–20 km [Pan *et al.*, 1991], and has been inferred as causing seismicity following the Upptyppingar dyke intrusion in central Iceland, close to the area we discuss below [Martens and White, 2013]. Other examples of seismicity in the lower crust thought to be triggered by volatile release have been reported from areas including Mammoth Lakes [Shelly and Hill, 2011], Lake Tahoe [Smith *et al.*, 2003] and the southern end of the Taupo rift in New Zealand [Reyners *et al.*, 2007].

In this paper we discuss seismicity from a series of swarms at 20 km depth beneath the Askja volcanic system (Figure 1), each of which show hundreds of repetitive, regularly spaced and identical seismic arrivals over short periods of typically 1–3 hours (Figure 2). The signal to noise ratio was good during the period these were recorded, enabling us to identify events under Vaðalda with local magnitudes (ML) as small as -0.5. We compare these earthquakes with a series of larger events in a cluster closer to the Askja central volcano which is about 10 km away but at a similar depth of ~20 km. The Askja cluster does not exhibit the same behavior as the Vaðalda swarms of multiply repeating events with identical waveforms, but their larger magnitude enables us to calculate more moment tensor solutions from the Askja earthquakes than from the Vaðalda earthquakes. We suggest that both the Askja and Vaðalda earthquake clusters were produced by similar processes of melt movement from sills in the lower crust.

2. Northern Volcanic Zone

Active volcanism in Iceland is concentrated along three volcanic rift zones: the Northern Volcanic Zone, the Eastern Volcanic Zone and the Western Volcanic Zone (Figure 1 inset). Each volcanic zone consists of a number of volcanic systems, which commonly comprise a large central volcano and an associated fissure swarm. The Northern and Eastern Volcanic Zones were generated less than 6 Ma ago by a ridge jump from an original position ~100 km further west [Jancin *et al.*, 1985; Garcia *et al.*, 2008]. These rift zones are the expression of the mid-Atlantic Ridge where it crosses Iceland, which in the region of our study is spreading at 19.6 mm/yr [DeMets *et al.*, 2010]. The mantle plume, which provides dynamic support and increased melt volumes, is centered beneath the area of thickest crust, under the Vatnajökull ice cap [Darbyshire *et al.*, 1998].

The Askja volcanic system is located at the southern end of the Northern Volcanic Zone. It is one of five en-echelon fissure swarms which make up the Northern Volcanic Zone. It has a large central volcano, called Askja, and a long fissure swarm which extends mostly to the north. To the east lies the Kverkfjöll volcanic system (Figure 1). Askja has been active in historical times with a large phreato-plinian rhyolitic eruption in 1875 which covered most of northeast Iceland in ash. Since this eruption there have been at least eight small basaltic fissure eruptions, with the last occurring in 1961 [*Global Volcanism Program*, 2015]. Geodetic measurements have been taken at Askja for five decades and show that the area is deflating and has subsided at least 25 cm since 1983 [*de Zeeuw-van Dalssen et al.*, 2013]. This is most likely due to a combination of deflation produced by the cooling of a shallow magma reservoir, and the rifting processes occurring due to plate spreading [*de Zeeuw-van Dalssen et al.*, 2012].

Vaðalda lies to the east of Askja volcano, between the Askja and Kverkfjöll volcanic systems (Figure 1). It is a monogenetic shield volcano formed during the last interglacial period (less than 0.8 Ma). It is not known whether the source of the lavas was directly below the shield volcano, or whether lava flowed laterally along a dyke from a more distant source, such as occurred with the 2014–2015 Holuhraun eruption 25 km to the southwest, which was fed from the Bárðarbunga central volcano a further 45 km to the south along the rift system [*Sigmundsson et al.*, 2015].

3. Data

In this paper we use the term cluster to denote deep seismicity in the mid lower crust that has been continually active over periods of many years. The term swarm refers to a short-lived (typically up to 3 hours or less) sequence of earthquakes within one of the clusters.

Data were collected from the Askja seismic network deployed by Cambridge University which has been active since 2006, supplemented by data from stations MKO, SVA and KRE operated by the Icelandic Meteorological Office. At the time of the Vaðalda swarms in December 2012 there were 35 seismometers active, most of which are shown on Figure 1, with a few at greater distances outside the map.

They comprised 28 x 30s Guralp 6TDs, 2 x 60s Guralp ESPCDs and 3 x 5s Lennartz short period instruments. Earthquakes from the very small ($ML < 1$), 20 km deep events discussed here were recorded on more than half of the network (typically 24 seismometers), which provides excellent depth control and azimuthal coverage: arrivals were recorded to offsets of 40 km, with a typical azimuthal gap of 70° .

During March 2010 when the cluster beneath Askja experienced a higher frequency of seismicity than normal, there were fewer instruments recording overall. However, because the deep events beneath Askja were larger than those in the Vaðalda swarms, most were recorded across the entire deployed array at the time of 22 seismometers.

4. Background Seismicity

Earthquakes in this region (Figure 3) are generated by three main mechanisms: a series of en-echelon faults around Herdubreifð which are the result of the differential motion between the Askja volcanic system and the Kverkfjöll volcanic system [Green *et al.*, 2014], an active geothermal region to the southeast of Öskjuvatn, a lake generated after a large caldera-forming rhyolitic eruption of Askja in 1875, and deep seismicity observed beneath the entire region first reported by Soosalu *et al.* [2010] and Key *et al.* [2011a,b]. The first two types occur at shallow depth (2–7 km), above a well defined brittle-ductile transition, the latter deep events are in the normally ductile lower crust at depths of 11–25 km.

The deep earthquakes form distinct clusters in four locations (Figure 3): the furthest east cluster (U, Figure 3) is from the 2007–8 Upptyppingar dike intrusion [White *et al.*, 2011, 2012; Hooper *et al.*, 2011; Martens and White, 2013] and is the result of an intruding and cooling dike; the second is beneath the shield volcano Kollótadyngja to the north of the area (K, Figure 3); the third beneath the shield volcano Vaðalda to the west of Upptyppingar (V, Figure 3); and finally 80% of the deep events in the region occur beneath Askja volcano itself (A, Figure 3). These deep microearthquakes all show clear P and S wave arrivals indicative of brittle failure. None of them show the low frequency events sometimes observed elsewhere in the deep crust and upper mantle [Okubo and Wolfe, 2008] and attributed to melt movement.

5. Analysis

5.1 Detection and location of seismicity

Earthquakes were automatically detected and initial locations generated using Coalescence Microseismic Mapping (CMM) [Drew *et al.*, 2013], which provided a catalogue of events. For events with a high signal to noise ratio arrival time picks were refined manually. More precise locations were then made by inputting event arrival times into the probabilistic location program NonLinLoc [Lomax, 2004], followed by double-difference relocations to improve their relative locations [Waldhauser and Ellsworth, 2000].

It is clear from the histogram of earthquakes binned in 10-day periods shown in Figure 4, which are generated from our CMM catalogue of events, that the clusters beneath Vaðalda and Askja have periods where the seismicity rate is much higher than the background rate (arrows, Figure 4). This is in contrast to the cluster beneath Kollóttadyngja, which has a more constant seismicity rate. In the case of the Vaðalda cluster during the single month of December 2012 more than 50% (87 CMM automatically located events) of the total seismicity over a period of five years was recorded. The peak rate excursions are less extreme in the Askja region because the background rate under Askja is much higher, even so, more than 70 events were detected in a 10 day period in March 2010 (arrow, Figure 4b).

The largest number of earthquakes beneath Vaðalda was recorded between the 23rd–27th December 2012, which is when the highly repetitive earthquakes occur (Figure 2). These are swarms lasting up to 3 hours, with a single earthquake mechanism repeating very frequently. Because the waveforms in individual swarms were so similar, we applied a further step of cross-correlation waveform detection [Shelly and Hill, 2011] to ensure that we had detected all the events in each swarm. Template waveforms were generated for each component from earthquakes with good signal to noise ratios, and accurate arrival times of both the P and S phases were generated by hand. The template waveforms were cross-correlated with the entire seismic trace for the period of 23rd–27th December. Peaks in the resulting cross-correlation function were taken as the arrival times of seismic phases. Note, that because of the short time window of the template waveform, arrivals need not be identical to be detected. Theoretical travel times from the average location of the

earthquakes for both P and S waves were then subtracted from the respective phase arrival times in order to produce potential origin times from every cross-correlated arrival time. Events were defined when more than 6 potential origin times occurred within 1 s of each other.

The potential events were then located using NonLinLoc [Lomax, 2004] with a 1D velocity model (Supplementary Table A) derived from nearby seismic refraction lines [Brandsdóttir *et al.*, 1997; Darbyshire *et al.*, 1998] and a local earthquake study by Mitchell *et al.* [2013]. Events which subsequently were not located within 5 km of the average swarm location were removed from the catalogue, along with events with a reported location error of more than 3 km. Arrival time picks with residuals of more than 0.5 s were also removed.

The resulting refined earthquake catalogue for the Vaðalda swarms consisted of more than 750 events. This is an order of magnitude more than it was possible to detect and locate using the automated CMM program. The primary reason for this was that the events were very small and so were only seen on about half of the seismic network, so did not have a sufficiently high coalescence value to be detected in the automated CMM catalogue.

During the March 2010 sequence under Askja, repeating earthquakes such as were seen during December 2012 beneath Vaðalda were not observed. Instead, short (< 10 min) periods of tremor were observed during particularly heavy periods of seismic activity, often following the larger events. The close association between the tremor and the locatable deep events suggests that it is from the same location and is related to the same processes. The tremor has the same frequency characteristics as recorded individual events, supporting the theory that they are located in a similar position. It also suggests that the periods of tremor are just periods of multiple earthquakes occurring closely in time and space, thus producing complex, overlapping codas.

5.2 Event Clustering

Beneath Vaðalda, earthquakes which are located closely in time within any given swarm are highly correlated, with correlation coefficients of 0.8 or higher (Supplementary Figures S3 - S10). Swarms were easily identified as periods of less than 3 hours in which the recorded events were highly correlated. The waveforms are highly similar on all stations, although the small size of the events leads to smaller than expected correlation coefficients and visible differences in the waveform due to noise (Figures 2 and 5). In total nine distinct swarms were identified (Table 1) although only six were large enough to be located and only two (swarms A and B, Table 1) had sufficient signal to noise ratios to allow further detailed analysis.

The largest earthquakes in an individual swarm have highly correlated codas out to more than 5 sec after the strong S wave arrival (Figure 2), indicating that all the events in a swarm have both the same source mechanism and a very similar location. The correlation coefficient matrix from station HOTT (Supplementary Figure S5) shows that at no point are two earthquake sources active at the same time, although often swarms have correlation coefficients of between 0.4 – 0.5 with each other, highlighting the small volume and similar mechanisms from which these microearthquakes are sourced.

5.3 Nature of Phase Arrivals

The larger events show clear P and S wave arrivals on nearby stations, but only S waves are seen from smaller events and on stations that are further away. On the closest station to the Vaðalda events, HOTT (see Figure 1 for location), which is located ~20 km directly above the earthquakes, spectra show that both high and low frequencies are present, with energy recorded from 2–20 Hz (Figure 6). Spectrograms (Supplementary Figures S1 and S2) show that most energy lies in the frequency range 5–10 Hz, which is typical also of small tectonic earthquakes in the brittle upper crust in this area. The predominance of S-waves on all but the closest stations from the deep events under Vaðalda is probably due to the higher amplitudes of S-waves compared to P-waves, and to the highly attenuating nature of the Icelandic crust in this volcanically active rift zone. Reported S-wave Quality Factors, Q_s , of 150–400 have been found in the lower crust of Iceland by Ólafsson *et al.* [1998], and as low as 140 by Menke *et al.* [1995]. The same authors report that the ratio of S-wave to P-wave Quality Factors is typically in the range 1.4–1.6. Attenuation probably increases markedly in the upper crust, which comprises multiple lava flows, rubbly horizons and ash layers. Recent

work by *Schuler et al.* [2013] shows that the effective Quality Factors for layered Tertiary basalts penetrated by boreholes are as low as 22–33 for P-waves and 13–17 for S-waves. These are typical of effective Quality Factors reported for other regions of layered basalts, and probably reflect scattering of energy rather than the intrinsic absorption [*Maresh et al.*, 2006; *Shaw et al.*, 2008]. Such low effective Quality Factors, even if only in the uppermost parts of the crust, serve to scatter the high frequencies from the waveforms and thus reduce the amplitude of the P-wave arrival beneath the noise threshold, while the larger amplitude S-wave arrivals may still remain visible above the noise.

5.4 Earthquake Magnitudes

Earthquake magnitudes were calculated using the local magnitude determination [*Bormann et al.*, 2013, *Richter*, 1935]. The formula has been calibrated against the magnitudes calculated and reported by the Icelandic Meteorological Office (IMO). The response due to the instruments was removed from the waveforms and the response of a Wood-Anderson seismograph convolved with them. For stations with a signal to noise ratio above a threshold of two, the largest peak-to-peak amplitude was picked automatically and inserted into the local magnitude formula. In addition, varying site specific effects have been accounted for by applying station corrections. The final magnitude is the mean of the magnitudes recorded on each component at each station. By using a multi-station approach the site and source effects are reduced and events within different clusters can be compared to each other. We use the method of *Aki* [1965] to calculate b-values, which show a good straight line fit (Figure 7). The magnitude of completeness is ML -0.17 and the b-value is 3.36 ± 0.17 , although because the magnitude range over which the b-value is calculated is quite restricted the errors are likely to be larger than stated here. Importantly the b-value could vary by an order of magnitude more than the stated error and still be higher than 1.

5.5 Magnitude distribution of Vaðalda earthquakes

Earthquakes within a swarm are not randomly distributed in time, but follow patterns in both inter-event times and magnitudes (Figure 7). The simplest case is swarm A (Table 1). This starts at 0455h on the 27th December and lasts for 2.5 hours (Figure 7a, b and c). Within this swarm 154 events occurred, ranging from

magnitude ML -0.5 to 0.1. At the beginning of the swarm events occurred every 10 s and have the smallest magnitude. At the end of the swarm events occurred every 100 s and reach their largest magnitude. The cumulative seismic moment release (black line, Figure 7b) shows that the rate of seismic moment release is not constant, but reduces over the period of the swarm. The seismic moment release ceases abruptly at the end of the swarm.

Other swarms of events show more complex behavior. Swarm B occurred from the 26th December at 2130h to 0000h on 27th December and contains the largest and clearest events of the entire episode (Figure 7d, e and f). This swarm contains events down to ML -0.4 and up to ML 0.4. Within the first 40 minutes of activity two cycles are seen of increasing, then decreasing amplitude. The events are at their most frequent, occurring every 8–25 s during these 40 minutes. During this time, the inter-event time is proportional to the size of the event, as in swarm A. The relationship between the inter-event time and the magnitude is not the same in both cycles (Figure 7d), suggesting there was either a small change in the frictional properties along the fault or in the stress field. After this, a third cycle starts which contains the largest events. It starts off the same as the previous cycles, with increasing inter-event time and increasing magnitude. Once the peak magnitude is reached the event size plateaus, while the inter-event time continues to increase.

Smaller magnitude earthquakes begin to be seen in swarm B between the larger earthquakes after 2220h (Figure 7f). These have high correlation coefficients with the larger events and so are members of the same family. At the same time the size of the larger earthquakes starts to decrease. During this period the seismic moment release rate is still decreasing as the small events do not account for the ‘missing’ seismic moment from the decrease in size of the large events. The larger events stop abruptly at 2300h, though the smaller events continue. The smaller events show a more random pattern of seismicity with no clear relationship between the inter-event time and the size of the earthquakes (Figure 7d and e). At approximately 2325h, the size of the events starts to decay. This behavior is not seen in most swarms, which tend to end abruptly. While the size of the events decreases, the inter-event time also increases. This is different from the pattern at the start of the swarm and shows a decoupling between the driving stress and the magnitude.

At 2355h a series of earthquakes occur which do not correlate with the other events. This series of micro-earthquakes consists of a single large event followed by a series of other small events which cannot be distinguished from each other. The large initial event locates in the area of interest, confirming that this is indeed an earthquake which is related to the Vaðalda cluster. Directly after this swarm ends, another swarm (swarm G, Table 1) starts with a different waveform, but is too small to locate accurately, so we cannot perform similar detailed analysis on it.

5.6 Relative relocations

Relative relocations of the earthquakes within each Vaðalda swarm were generated by calculating highly accurate, subsample, differential travel times using a cross-correlation technique. Waveforms were filtered using a bandpass filter from 2–20 Hz, then cross-correlated with each other in a window from 0.5 s before the theoretical arrival time to 3 s after it using the GISMO suite of computer programs [Reyes and West, 2011]. The differential times were combined with the catalogue arrival time picks from the cross-correlation detector and input into HypoDD [Waldhauser and Ellsworth, 2000] (Figure 8). The gain in accuracy is achieved by relocating the earthquakes with respect to each other, thereby reducing the impact of velocity heterogeneities along the predominantly shared ray paths. The resulting earthquake locations have horizontal and vertical uncertainties reported by HypoDD of less than 10 m.

To further assess the errors on location, jackknife tests were performed by repeating the HypoDD analysis but removing one station at a time. Results depend on which swarm is being analyzed but show that the true error on a single event varies from between 8 m in all directions in swarm B to more than 150 m horizontally and 300 m vertically in other swarms (Supplementary Figures S15–S19). The high cross-correlation coefficients between events, especially later in the coda, imply that the earthquakes in any given swarm come from a region that is smaller than the first Fresnel zone (~ 325 m at 5 Hz). This is significantly larger than the scatter observed for some swarms (Figure 8), suggesting that there is another cause for the scatter in locations that we observe for swarms A, C, D, E and F.

Swarm B, which is by far the clearest, has the smallest errors and has more than 140 events clustered within a sphere less than 20 m across, significantly smaller than the error in location implied by the waveform similarity argument. This is because the time difference between the P and S wave arrivals is identical for all stations across the network (Supplementary Figures S11–S22). The events in Swarm A, which have lower signal to noise ratios and are therefore less well correlated with each other, also show no apparent lag in the P to S wave arrival times at many stations (Supplementary Figures S11–S22) indicating that they too are likely to come from a single location. The spread in locations is therefore entirely due to picking errors, which are higher in all the swarms except Swarm B because of low signal to noise ratios. Because of this, we propose that in fact all the swarms probably originate from single points located approximately in the centroid of each swarm (Table 1). This would explain the high degree of similarity between the events within a swarm and suggests that the errors for each swarm vary between 20–300 m horizontally and 20–500 m vertically (Table 1).

5.7 Moment tensor solutions

It was not possible to determine the polarities of P wave arrivals for most individual earthquakes in the Vařalda swarms because the P-waves contain predominantly high frequency energy that is attenuated strongly in its passage through the crust. However, because the waveforms from many events are almost identical (Figures 2, 5 and Supplementary Figures S11–S14), we are able to stack the traces together to improve their signal to noise ratio. This resulted in an increase in the reliability of measured polarities. We also added constraints from the polarities of S-H wave arrivals picked on the transverse component, which were generally recorded with much larger amplitudes on our seismometers and greatly added to the reliability of the solutions. Moment tensors were derived by a Bayesian inversion of the polarities [Pugh, 2015] taking into account the probability of an incorrect polarity pick and a 1 km uncertainty in location. In Figure 9a we show the resultant moment tensor solution for earthquakes in swarm A, and in Figure 9d from swarm B beneath Vařalda. Neither can be fitted with double couple solutions, but require a volumetric component. On Hudson type plots [Hudson *et al.*, 1989], the most probable solution (shown by red stars on Figures 9a and 9d) lies between the double-couple solution (labeled DC) and an opening tensile crack (labeled TC+). The full scatter in possible solutions is shown by the blue/green shading.

Similar activity under Askja in March 2010 events contained larger events with better signal to noise ratios, so we were able to make moment tensor solutions from individual events. We found two unambiguous events which again required volumetric increases as well as shearing (large moment tensor solutions on Figure 10), while the remaining events from March 2010 could be fitted by double-couple solutions (red dots and smaller moment-tensor solutions in top left part of Figure 10. Figure 11 shows the moment tensor solution from one of the events (23:08 3/1/2010), together with the waveforms of the P-wave arrivals; all but two of the arrivals are positive, and the focal sphere is well covered by the observations. As for the Vaðalda events, the possible solutions for this Askja event suggests an opening crack on the Hudson plot (Figure 14a).

6. Discussion

6.1 Induced seismicity in the lower crust

Earthquakes are not commonly seen in the lower crust of Iceland because the long-term driving stresses are continually released by the ductile nature of the crust at these depths. Commonly during the intrusion of melt into the crust, earthquakes are seen near the tip of the propagating melt front [*Rubin and Gillard, 1998; Segall et al., 2013; Sigmundsson et al., 2014*], where stresses are maximised. In the earthquakes observed beneath Vaðalda and Askja, we do not see any systematic spatial and temporal migration that would be expected from such a model, as the earthquakes do not move in a consistent direction with time (although migration over distances of less than our minimum resolution of ~20 m would not be resolved). Furthermore, earthquakes during a phase when the intrusion is increasing in size would not be likely to exhibit highly similar mechanisms because the position where the rocks were fracturing would be different every time.

Brittle failure can also be induced within the intrusion itself, where during periods of low melt supply the intrusion freezes, at least at its margins. Basaltic glass reaches a peak in strength just after it cools through the glass transition [*White et al., 2012*] and can then fail in a brittle manner when the magma pressure behind the frozen plug reaches a critical threshold. In the Upptyppingar dike intrusion, failure occurs on both the upper and lower edge of the intrusion resulting in opposing focal mechanisms (*White et al., 2011*). However,

in our Vaðalda data the earthquakes occur much too frequently (sometimes every 10s) for the creation of a substantial melt plug by freezing between earthquakes.

In order to induce seismicity in normally ductile materials the strain rate must be high and/or the pore-fluid pressure must increase. This can happen in volcanic regions where the movement of melt within the crust must occur for melts to migrate from the mantle, where it is formed, to the upper crust where it is either erupted or intruded. Additionally, CO₂ within the melt is saturated at depths of 15–20 km and so degasses as a batch of melt ascends through the crust or as it freezes. Both the melt and exsolved CO₂ provide fluids which may increase the pore fluid pressure. High b-values have been associated with many processes including high thermal gradients adjacent to melt bodies, the presence of fluids, and deformation in a ductile regime [Scholtz, 1968; El-Isa and Eaton, 2014]. We observe b-values greater than one from the Vaðalda swarms indicating that in the immediate area there could be fluids and high thermal gradients, either of which could induce seismicity in a normally ductile regime.

The repeating nature of the Vaðalda events within each swarm indicates that the mechanism generating the seismicity is non-destructive. Similar processes are likely to occur during the periods of tremor in March 2010, although the non-destructive nature of the source is difficult to determine given the small time gap between the tremor events.

6.2 Conceptual model

At the tips of intrusions the stresses are maximised [Hill, 1977] and therefore this is where earthquakes are most likely to occur [Toda *et al.*, 2002], producing typical ‘dog-bone’ like geometry. We propose that the observed microseismicity under Vaðalda occurs on small high angle thrust faults at the tips of sill intrusions in the lower crust (Figure 13). If we assume that all the events in a single swarm are located at a single point then it is likely that a single fault is active at a time. When melt is injected to form a sill, a high stressing rate, along with the injection of small amounts of fluid (either melt, CO₂ or both) along a fault may induce stick-slip motion on the fault. Because the swarms beneath Vaðalda are located up to 1 km from each other (Figure 8), we propose that each swarm observed is caused by melt movement from different sills, in a

heavily intruded region of the Icelandic crust. The Askja earthquakes during March 2010, despite not having the repeating earthquakes seen in the Vaðalda swarms, are still likely to be due to stick-slip motion on faults induced by the intrusion of melt into the area. The lack of highly similar events under Askja suggests that in this case the seismicity was a destructive mechanism. The short periods of tremor could be due to similar processes to the Vaðalda swarms. In this case the tremor is due to many events repeating, just very quickly and not at regular intervals.

The emplacement of sills into the crust is controlled by the presence of a rigidity contrast [*Kavanagh et al.*, 2006] and specific thermal and dynamic parameters [*Chanceaux and Menand*, 2014]. From long term monitoring of the Askja area we know that this region has been active seismically for at least 8 years and that conditions must be favorable for the formation of sills [*Soosalu et al.* 2010; *Key et al.* 2011a,b]. Causes of the rigidity contrast cannot be known, but the existence of cooled intrusions in the lower crust may cause contrasting rock properties which could act as mechanical barriers. This could form regions of repeated sill intrusion causing a small volume to be active similar to what is seen presently in the deep clusters of earthquakes beneath the Askja volcanic system.

Calculated moment tensors reveal that all the earthquakes in the Vaðalda swarms and some of the earthquakes in the March 2010 below Askja require a volumetric component to fit the polarity observations. The best fitting solutions lie between the double-couple point at the center of a Hudson plot and an opening tensile crack. Any failure of frozen melt on the sides of a sill would be expected to be parallel to the intrusion rather than at a high angle to it. There are some double-couple solutions in the Askja cluster that have one sub-horizontal nodal plane and one high-angle plane (Figure 10), which might therefore be caused by fracture of melt frozen on the margins of a sill. However, the non-double-couple solutions all require high-angle fault planes with a volumetric increase. We interpret this as caused by melt intrusion from the margin of a sill either into the country rock or into an opening crack connecting one sill to another slightly shallower one. The high-pressure fluids injected along such a crack could also reduce the normal stress across the crack and thus promote further failure on the same crack with the same orientation; this would explain the highly repetitive waveforms we observe in the swarms under Vaðalda.

6.3 Magnitude Distribution

Cluster A showed a consistent variation in the inter-event time as the cluster proceeded. During some volcanic eruptions, the inter-event time decreases up to the time of the eruption before merging into harmonic tremor [Hotovec-Ellis *et al.*, 2013]. This can be modeled within a rate and state framework as an increase in the stressing rate [Dmitrieva *et al.*, 2013]. In contrast, the behavior of the events in the Vařalda swarm A (Figure 7b) show an increase in inter-event time and so can be thought of as resulting from a decreasing stressing rate. Formation of a new sill will induce local stresses on the surrounding country rock, either forming new faults or reactivating old ones. In addition to this, the injection of fresh melt will cause an increase in the pore fluid pressure along faults around the sill. Melt or exsolved CO₂ from decompression or partial crystallisation would be injected along these faults and favor the deformation of the country rock by brittle rather than ductile means. As the melt supply slows, the stressing rate is reduced, resulting in longer time intervals between events and a reduction in the seismic moment release rate. The seismicity could end if either, the stressing rate becomes too low to induce brittle failure or a reduction in the pore fluid pressure results in an inability to cause brittle failure. In reality both mechanisms are likely to operate. The small size of the events indicates that the fluids which induce brittle failure cannot intrude far from the parent sill itself.

Studies of microearthquakes [Abercrombie, 1995; Ide *et al.*, 2003] have shown that the fault scaling relations derived for larger earthquakes apply down to at least Mw 0. Assuming that the bulk modulus of the lower crust is 75 GPa [Auriac *et al.*, 2013], the Poissons ratio is 0.25 [Auriac *et al.*, 2013], the ratio between fault length and displacement is 5×10^{-5} and that local magnitude can be converted to moment magnitude, M_0 , by the empirical relationship, $ML = Mw = (2/3)\log_{10}(M_0) - 6.07$ [Borman *et al.*, 2013], we estimate a fault radius of 10 m for the largest earthquake we observe, which has ML 0.4. Observed variations in the stress drop during earthquakes results in a factor of two variation in the size of the calculated fault plane. The size of the calculated fault plane is less than the minimum resolution in the position of the located earthquakes (20 m), indicating that we should not expect to see migration of activity on the scale of a fault and therefore the different swarms are caused by different intrusions rather than one repeated intrusion.

The size of the fault plane is partially determined by the distance fluids can percolate along the fault plane adjacent to the intrusion, as the presence of fluids fundamentally controls whether or not brittle deformation occurs. The damage zone around dikes is observed to be approximately one dike width away from the edge of the dike [Kavanagh, 2011]. This relates the size of the earthquake to the thickness of an intrusion. From the observed earthquake magnitudes we infer that they were caused by a small intrusion, on the order of ~10 m thick. Using length to thickness scaling relations calculated for sills [Cruden and McCaffrey, 2002; McCaffrey and Cruden, 2002; Menand, 2008] a thickness of 10 m results in a sill with a radius of ~1000 m. A sill of this size would take 0.5 years to cool in the lower crust [Turcotte and Schubert, 2002], so the short duration of the swarms implies that the sill freezing is not the cause of the abrupt end to the seismicity. It is more likely that the stressing rate in the sill induced by the injection of melt slowed so that brittle failure was no longer possible. The presence of multiple crustal intrusions in this area, which are unlikely to have cooled back to ambient temperature, create large impedance contrast scatterers in the region close to the microearthquake sources. These are most likely to be the cause of the local scattering observed in the seismic waveforms.

In contrast to swarm A, swarm B shows a much more complex earthquake sequence and displays how the interplay between the stressing rate and the presence of fluids can become very complex. The seismicity at the beginning of the swarm shows similar features to swarm A, with increasing inter-event times and increasing magnitudes. However, the size of the earthquakes saturate ~40 minutes after the start. This is likely to be due to the finite distance that fluids can migrate and cause the required high pore fluid pressures for a fault to fail. When the smaller earthquakes start, they do not fully account for the ‘missing’ moment release from the reduction in size of the larger events. This could be the result of a reduction in the spatial distribution of pore fluid pressure around the fault. When different areas of the fault have the pore fluid pressure reduced below a critical value, the area available to slip seismically would reduce, thus reducing the magnitude of events. This, in combination with a reduction in the stressing rate could result in a reduction in the rate of seismic moment release.

7. Conclusions

During December 2012 several swarms of small earthquakes were observed at 20 km depth in the lower crust beneath Vaðalda. Each swarm lasted for less than 3 hours and contained events ranging from ML -0.5 to 0.4 with near-identical waveforms repeating up to every 8 s. We have shown that the apparent spread of the locations is likely to be due to errors in the picked arrival times and that the events from a single swarm come from a single point. Because each swarm is located in a unique position our preferred model is that they were caused by a series of small sills intruded into the lower crust beneath Vaðalda, an area which has been persistently active since 2006. Similar events were observed at the same depth beneath Askja volcano, some 10 km away, although in the latter case without multiply repeating near-identical microearthquakes. Due to the small size of the intrusions, there were no observable changes in elevation at the surface. Seismicity was triggered on small faults located at the tips of the intrusions in the normally ductile lower crust through a combination of the high strain rates induced on the country rock and high pore fluid pressures induced by the movement of fluids, either melt or CO₂, along fault planes. The faults may either accommodate sill growth, or may be pathways between sills along which melt is migrating from one sill to another (Figure 13). The small magnitude of the earthquakes show that the active fault plane must have been limited in size.

Moment tensors calculated for the clusters beneath Vaðalda and for two of the events recorded in March 2010 under Askja cannot be fitted with double-couple mechanisms but instead require an opening tensile crack component. We propose that an opening crack is required in order to create the space to intrude melt into the lower crust. This is in contrast to upper crustal intrusions which are always double-couple sources. Although a few of the moment tensor solutions from the March 2010 Askja cluster show sub-horizontal fault planes, as might be expected from failure parallel to a sill, the bulk of the mechanisms, including the non-double-couple solutions show high angle fault planes. High stresses induced at the tips of intrusions are likely to facilitate failure at ~45° to the sill.

Fault planes such as these could behave as ‘highways’ for melt within the mid to lower crust and form a distributed network of linked sills in the lower crust transporting melt from the mantle to either form the lower crust or eventually to erupt at the surface.

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Tables

Table 1 - Vaðalda swarm information

Swarm	Start Time	End Time	Number of Events	Centroid Location			Error		
				Latitude	Longitude	Depth / km	x / m	y / m	z / m
A	12-27-2012 04:55	12-27-2012 07:30	154	64.0335	-16.4847	20.8	150	150	500
B	12-26-2012 21:30	12-27-2012 00:00	144	65.0312	-16.4921	20.937	20	20	20
C	12-23-2012 11:30	12-23-2012 12:22	15	65.0299	-16.4828	21.3	100	100	500
D	12-23-2012 12:22	12-23-2012 13:40	117	65.0357	-16.4696	21.8	400	300	300
E	12-23-2012 19:25	12-23-2012 20:10	84	65.0323	-16.4782	21.21	150	300	200
F	12-24-2012 01:00	12-24-2012 06:00	74	65.0342	-16.4787	21.3	100	100	300
G ^a	12-27-2012 00:00	12-27-2012 01:00	92	-	-	-	-	-	-
H ^a	12-27-2012 02:20	12-27-2012 03:00	42	-	-	-	-	-	-
I ^a	12-27-2012 03:30	12-27-2012 03:55	8	-	-	-	-	-	-

^a Swarms G, H and I did not produce stable locations in HypoDD and so have no location information

Figure Captions

Figure 1. Map of the study area. Stations which detected the micro-earthquakes under Vaðalda are shown as filled yellow triangles, other deployed stations as empty triangles. Stations HOTT, MKO, ASK, FLAT, HRUR, MIDF, KATT and LIND referred to in the text are labeled. The shield volcano Vaðalda, the table mountain Herðubreið, the central volcano Askja and the caldera lake (Öskjuvatn) formed by the most recent large eruption (1875) are labeled in white. Green and red stars show the location of micro-earthquake swarms from 2010 (Askja) and 2012 (Vaðalda) respectively, which are discussed in this paper. The Askja and Kverkfjöll Volcanic systems referred to in the text are labeled in white, and aligned with the rift axes. Lineated black lines show the mapped fractures of Hjartardóttir et al. [2009, 2011] and delineate the active rifts. The black box shows the area displayed in Figure 3. Inset shows the location of the study area in Iceland (black box) along with the three neo-volcanic zones: Western Volcanic Zone (WVZ), Eastern Volcanic Zone (EVZ) and the Northern Volcanic Zone (NVZ).

Figure 2. Upper panel shows an enlargement of a selection of earthquakes from two different clusters aligned on the S arrival (0.5 s). Lower panel shows the repetitive earthquakes recorded on the east component of station HOTT on 26th December 2012.

Figure 3. Background seismicity from double-difference relocations in the vicinity of Askja from 2006 to present. Blue earthquakes are in the brittle zone of the upper crust, orange earthquakes in the normally ductile lower crust. Contour lines around the lower crustal events are drawn at 2, 20, 50 and 100 earthquakes per km² and indicate the high density of these events. Map shows locations of the cross-sections displayed on the right. Yellow boxes surround the events located below the brittle-ductile boundary and used for the histograms in Figure 4: A, Askja; K, Kóllottadyngja; U, Upptyppingar; V, Vaðalda.

Figure 4. Ten- day histograms for events within the lower crustal clusters defined in Figure 3. a) Vaðalda; b) Askja; and c) Kollóttadyngja. The black line shows the cumulative seismicity. Arrows mark the swarms discussed in this paper.

Figure 5: Stacked waveforms from Vaðalda swarm B for the stations with locations labeled in Figure 1. The light grey region shows the maxima and minima of the stack, the dark grey region shows one standard deviation and the black line shows the stacked waveform.

Figure 6. Normalized spectral content for each component of stations HOTT (black) and LIND (red) (see Figure 1 for locations).

Figure 7. Recorded microseismicity from swarms A and B. a), b) and c) show swarm A and d), e) and f) show swarm B. Points are colored consistently across all panels according to the time of day in panels c) and f) for swarm A and B respectively. Panels a) and d) show the relationship between inter-event time and the local magnitude. In panels b) and e) the colored dots display the inter-event time vs time of day and the black line shows the normalized cumulative moment with time. Panels c) and f) show the recorded waveform on the east component from station HOTT (see Figure 1 for location). The magnitudes of the microearthquakes that were located are indicated by colored dots.

Figure 8. Relocated seismicity using HypoDD [Waldhauser and Ellsworth, 2000]. Top figure shows map view and each swarm is colored: A - blue, B – red, C - black, D - green, E - light blue and F - purple (see legend). For description of swarms see Table 1. Contours show the density of earthquakes in each cluster and are drawn at values of 500, 2,500, 5,000 and 10,000 events per km². Bottom panels show cross sections orientated east-west (left) and north-south (right) colored and contoured the same as the map view. The aspect ratio is 1:1.

Figure 9: Moment tensor solution for swarms A and B. a) and d) show the probability density functions (blue=lower probability, yellow=higher probability) of the solutions considering error in both location and polarity on a Hudson type plot (Hudson et al., 1989) for swarm A and B respectively. Red star indicates the maximum likelihood solution in each swarm. b) and c) show the observed P and SH polarities respectively and the final maximum likelihood solution for swarm A. e) and f) show the observed P and SH polarities respectively and the final maximum likelihood solution for swarm B. DC is double-couple solution, TC+ and TC- are opening and closing tensile cracks respectively, CLVD+ and CLVD- are positive and negative compensated linear vector dipole. Upward facing red triangles show compressional arrivals, downward facing blue triangles show dilatational arrivals. Symbols are scaled so that the size is equivalent to a 1 km error in location in east-west, north-south and vertical directions.

Figure 10: Upper panel shows a map view of the deep earthquakes (grey) beneath Askja. Earthquakes recorded during March 2010 are colored red. 16 maximum likelihood double-couple solutions from this period are indicated by the beachballs on the left. The two events which were of sufficient quality to perform a full moment tensor solution are shown by the larger beachballs on the right and location colored in blue. Lower panel shows the cross-section along the line indicated in the top panel. Colors are the same as the top panel.

Figure 11: Calculated moment tensor for the event at 23:08 3/1/2010. a) shows the probability density function (blue=lower probability, green=higher probability) of the solution considering errors in both location and polarity on a Hudson type plot (Hudson et al., 1989). Red star indicates the maximum

likelihood solution. DC is double-couple solution, TC+ and TC- are opening and closing tensile cracks respectively, CLVD+ and CLVD- are positive and negative compensated linear vector dipole. b) shows the observed P-wave polarities with the maximum likelihood solution. Upward facing red triangles show compressional arrivals, downward facing blue triangles show dilatational arrivals. The recorded waveforms at all stations are shown around panel b). The red/blue line indicate the time of the arrival for compressional/dilatational arrivals respectively.

Figure 12. Cartoon showing schematic distribution of sills in the region recorded by the observed microearthquakes. The change in shear stress expected on faults by the inflation of a sill (inset) is shown on the top sill. Red and yellow colors in the top diagram indicate an increase in the shear stress [Toda *et al.*, 2011]. Earthquakes are generated on faults brought to brittle failure by the combination of a high strain rate and the presence of fluids within the fault. Faults such as this could provide links for melt movement between sills in the lower crust although have not been directly imaged in this work. An example opening-crack moment tensor from swarm B is shown in the lower right corner.